Test of models for the Cascadia great earthquake rupture area using coastal subsidence estimates for the 1700 earthquake

Annual Project Summary

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Investigations Undertaken

Seismic hazard assessments for Cascadia great earthquakes are largely based on the rupture area and slip displacement predictions of dislocation models constrained by present interseismic geodetic data and geothermal data. In this study, we examine the consistency of such models with compiled coastal coseismic subsidence estimated from paleoseismic studies. We have now completed the first phase of this investigation, comparing the predictions of simple elastic dislocation models constrained by geodetic data, with coastal data for the most recent great event in 1700. The results have recently been published (Leonard et al., 2004). Although the uncertainties in coastal subsidence are substantial, the results provide good support that the dislocation models provide a robust first order prediction of the rupture area and displacement. The two discrepancies are at the north end of the subduction zone off Vancouver Island, and the south end off N. California, where the predicted subsidence is much greater than that observed. The great earthquake cycle appears to be elastic to a good approximation. In the next phase, we extend the study in two ways.

Brief statement of the Investigations

In our continuing study: (1) We examine the effect of varying the rupture width, transient rupture behavior, and viscoelastic earthquake relaxation on the predicted coastal coseismic and postseismic subsidence . (2) We review, tabulate and summarize the coseismic subsidence data for great earthquakes prior to the 1700 event, and compare the results to the predictions of the dislocation rupture models.

Non-technical Summary

Seismic hazard assessments for Cascadia great earthquakes are largely based on the rupture area and rupture slip predictions of theoretical models. The models are constrained mainly by the pattern of current ongoing horizontal and vertical deformation in the coastal area from precision GPS and other geodetic data. There also are thermal constraints which give a similar result. In this study, we examine the consistency of such models with compiled coastal coseismic subsidence estimated from paleoseismic studies, i.e., does the model-predicted coastal coseismic subsidence agree with the abrupt subsidence that actually occurred in the most recent

event in 1700? The results provide a good first order confirmation of the theoretical models, except that the coastal subsidence and inferred rupture displacement appear to be smaller than predicted at the north end of the subduction zone off Vancouver Island, and the south end off N. California. We have completed the first phase of this investigation and the results have recently been published (Leonard et al., 2004). In the next phase of the study we now extend the study to deal with more complex theoretical models and with the coastal subsidence in the great events prior to the 1700 earthquake. Study of the older events will allow us to estimate if the 1700 event is typical of great earthquakes on this margin.

Reports Published

Leonard, L.J., R.D. Hyndman, and S. Mazzotti, 2004, Coseismic subsidence in the 1700 great Cascadia earthquake: Coastal estimates versus elastic dislocation models: Geological Society of America Bulletin, v. 116, no.5/6, p. 655-670.

A compilation of coseismic subsidence estimates is available at the Geological Society of America Data Repository, Item 2004076, "Depth of 1700 horizon and estimated 1700 Cascadia coseismic subsidence", at Web address: http://www.geosociety.org/pubs/ft2004.htm

Results

1. Introduction

Elastic dislocation models of megathrust earthquake rupture for the Cascadia subduction zone predict that the coast of southwest Canada and northwest US will subside abruptly during a great earthquake. These models are consistent with widespread evidence for coastal subsidence following the 1700 Cascadia megathrust earthquake, e.g., submerged trees and buried marshes. In our initial study now completed, Leonard et al. (2004) showed that the estimates of subsidence from marsh paleoelevation studies (0.3-2.0 m) are generally consistent with the subsidence predicted using a simple elastic dislocation model. For much of the Cascadia margin, the amount of fault slip in the models (10-50 m) is in agreement with the expected strain accumulation along a locked fault for a 500-800 yr earthquake recurrence interval. We follow up on this work to examine: (1) the sensitivity of model results to variations in the earthquake rupture width, (2) possible reasons for the observed small amount of subsidence at the north and south ends of the subduction zone, and (3) the effects of great earthquake post-seismic slip and viscoelastic relaxation. (4) We also assess whether the 1700 great earthquake was typical of previous events recorded at Cascadia. In most Cascadia marshes, the 1700 buried soil is the uppermost of a sequence of up to 11 buried soils; we compile estimates of coseismic subsidence for these pre-1700 events for a comparison with the 1700 record.

2. Variations in Earthquake Rupture Width

In their models of Cascadia earthquake rupture, Leonard et al. (2004) used a rupture zone that is consistent with the seismogenic zone inferred from interseismic geodetic observations and thermal data (e.g., Hyndman and Wang, 1995). The seismogenic zone is defined as the part of the fault that is fully locked between earthquakes. During the earthquake, it is assumed that the entire seismogenic zone ruptures, releasing the amount of strain accumulated since the previous earthquake. In the simple model, there is a transition zone downdip of the rupture zone, where fault slip linearly decreases to zero. Leonard et al. (2004) used a transition zone width similar to that for interseismic geodetic models that assume a linear change from full locking to free slip (e.g., Hyndman and Wang, 1995). Note that there is some question about how the interseismic transition zone relates to the coseismic fault behaviour (Wang et al., 2003).

We first have examined the sensitivity of coseismic vertical displacement to the width of the rupture zone, to evaluate the degree of constraint from the coastal data. Following previous studies, we assume that full rupture extends to the deformation front, so the downdip limit is the main variable. The reference downdip limit of full rupture is taken as the downdip limit of the seismogenic zone from thermal/geodetic studies, and we have varied this limit so the width of the rupture zone varies by +/-25%. This range is consistent with the uncertainties in the thermal/geodetic seismogenic zone (Hyndman and Wang, 1995). The downdip limit for the transition zone is that used by Leonard et al. (2004). Variations in this point have little effect on surface deformation because of the small amount of slip near the downdip limit; surface deformation is primarily affected by slip along the shallower part of the fault. In our models, the amount of slip in the full rupture zone is equal to the accumulated strain on the fault due to plate convergence over 800 yrs, the length of time between the 1700 event and the previous earthquake (Leonard et al., 2004 and references therein).

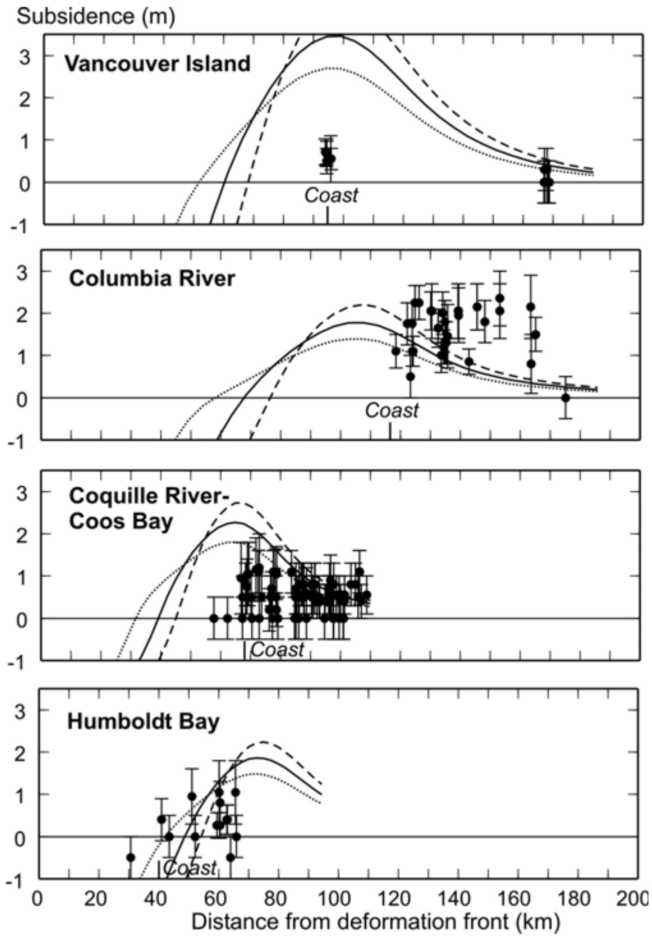

Figure 1. **Effect of variations in the rupture width on the modelled co-seismic surface vertical motion (dashed +25%, dotted -25%). A 800 yr. strain accumulation is assumed. Marsh subsidence data are within 75 km of each profile.**

Figure 1 shows the effect of variations in the rupture zone width for four profiles perpendicular to the margin (see **Fig. 4** for locations, and Leonard et al., 2004, for data details). An increase in the rupture width increases the magnitude of vertical deformation of the surface. In addition, the point where deformation changes from uplift near the deformation front to subsidence inland approximately coincides with the downdip limit of full rupture; a wider rupture zone shifts this point landward. For most of the margin, a variation of 25% in the rupture width has little effect on the amount of modelled coastal subsidence. Discrepancies between the marsh observations and dislocation model can be resolved by varying the amount of fault slip (Leonard et al., 2004), suggesting that the coseismic rupture zone is consistent with the seismogenic zone inferred from thermal and interseismic geodetic data. In order to constrain the rupture width from coseismic subsidence data, it is necessary to have data that lie near the uplift/subsidence transition point. The transition point occurs offshore for most of the Cascadia margin, so the coastal marsh data provide only a general constraint to the rupture width. Along the coast, the amount of subsidence in the dislocation model depends on both the rupture width and the amount of slip; an increase in either produces a greater subsidence.

One area where the downdip rupture limit may be better constrained from marsh data is near the Humboldt Bay, California, profile. Here, marsh data indicate only a small amount of subsidence and possibly some uplift, suggesting that the downdip limit of rupture underlies the coast. The data may be better fit using a rupture zone that is slightly narrower (~25%) than the model seismogenic zone (**Figure 1**). However, even with a smaller rupture width, the low amplitude of surface deformation suggests that slip in this area may have been less than that expected for 800 yrs of strain accumulation (see section 3.2).

3. North and South Ends of the Subduction Zone

3.1 Vancouver Island region

In the comparison of dislocation models with marsh subsidence data by Leonard et al. (2004), the north end of the Cascadia subduction zone on coast of Vancouver Island showed much less coseismic subsidence from marsh data than predicted. Coseismic subsidence estimates are clustered in two groups at 95 km and 170 km from the deformation front. The data closest to the deformation front indicate only 0.6 m of subsidence, much less than the over 2 m predicted for a fault slip of 36 m for accumulated strain over 800 years (**Figure 2**). One way to match the observation is by decreasing the width of the rupture zone to less than 50% of the thermal/geodetic model seismogenic zone, i.e., all well offshore (**Figure 2**). This model is seriously inconsistent with interseismic geodetic constraints on the currently locked zone. The second way to match the observations is to postulate the full width of the model locked zone to have ruptured, but with less rupture displacement than expected for the accumulated strain in 800 years. A good fit to the observations is found using a slip of less than 10 m (Leonard et al., 2004, **Figure 2**).

If it is assumed that the 1700 earthquake released all the accumulated strain on the fault, and there is no aseismic slip, a rupture displacement of 10 m suggests thrust earthquakes in this region that are independent of the M~9 events for the central region. An earthquake recurrence rate for independent earthquakes on this segment of the margin is then ~200 years, given that the plate convergence velocity is ~4.5 cm/yr. The recurrence rate for central part of the margin is 500-600 yrs (Goldfinger et al., 2003; Leonard et al., 2004, and references therein), and there is no evidence for a higher recurrence rate in the offshore turbidites of the northern area (Goldfinger et al., 2003). In addition, the expected moment magnitude for a megathrust earthquake along the northern 100-200 km of the Cascadia subduction zone is 8.2-8.4 (assuming a rupture width of 75 km and 10 m of co-seismic slip). It is unlikely that there has been an undetected earthquake with a magnitude greater than 8 along this part of the margin in the last 300 yrs (about time of written history). Therefore, if the recurrence is indeed ~200 years, such an earthquake is overdue. Other alternatives are: (1) a large part of the convergence is accommodated by aseismic slip, perhaps following the earthquake (over timescales longer than recorded by the marsh sediments), (2) the thrust is fully locked, but the 1700 event had especially small rupture in this region, with larger than average rupture in some previous events. This explanation also

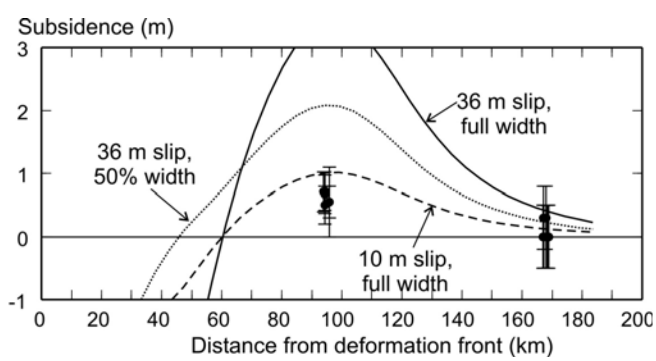

Figure 2. Variations in the amount of slip and rupture width for the Vancouver Island profile.

suggests that future events may have larger than average rupture. Unfortunately, there is only one buried marsh on this part of the margin (perhaps because of postglacial rebound or tectonic uplift) so the subsidence in previous events is not constrained. (3) The marsh data may include the effects of post-seismic fault motion, such that the maximum amount of coseismic subsidence is not recorded in the marsh data. This is discussed in greater detail below.

An intriguing observation about the Vancouver Island part of the Cascadia margin is that the 1700 earthquake is the only earthquake preserved in the coastal marshes above older bedrock or Pleistocene sediments (Leonard et al., 2004); further south, multiple earthquakes are recorded in the buried soils. One explanation is that there is a long-term tectonic uplift of Vancouver Island at a rate faster than sea level rise (~1.8 mm/yr), such that the previous earthquake records have been erased by erosion. The uplift mechanism is unclear; a study by Clague and James (2002) suggests only small post-glacial rebound in this area. The lack of preservation of previous earthquake signals, as well as the small amount of subsidence in the 1700 earthquake make the Vancouver Island region a key area for future work.

3.2 Gorda Plate region

The coseismic marsh data at the southern part of the Cascadia margin (south of ~43°N) suggest earthquake slip of less than 20 m in the 1700 earthquake, about half of that expected for 800 years of strain accumulation (Leonard et al., 2004, **Figure 1**). As in the Vancouver Island region, it is possible that the 1700 earthquake did not release the full amount of accumulated strain or that there is a higher earthquake recurrence rate (200-300 yrs) in this region than the rest of the margin. This second explanation is supported by the observation that there are a larger number of offshore turbidites at the southern end of the Cascadia margin compared to the central region (Goldfinger et al., 2003). However, the tectonics of the southern Cascadia margin also are complicated by the proximity of the Mendocino Triple junction and breakup of the Gorda Plate (e.g., discussion in Leonard et al., 2004). In this area, there are a number of possible fault sources for large earthquakes, and thus, the offshore turbidites may record earthquakes other than megathrust earthquakes. In addition, the complex tectonics in the vicinity of a triple junction mean that the elastic dislocation model for megathrust earthquake rupture should be interpreted with caution in this area (Wang et al., 2003; Wang, 2004).

4. Transient Fault Behaviour

The elevation differences inferred from marsh studies represent the change in elevation between the time of the earthquake and the time when the marsh sediments (with elevation indicators) were deposited following the earthquake. Several studies have suggested that the elevation indicators are re-established within a few weeks of an earthquake (see summary by Leonard et al., 2004, and references therein), but this timing is not well-constrained. Thus, the elevation differences recorded in marshes may include slow vertical ground motion due to processes following the earthquake rupture. In this section, we consider to processes that may affect the vertical deformation of the coastal region, (1) post-seismic slip on the deep part of the fault, and (2) viscoelastic stress relaxation.

Post-seismic slip. A megathrust earthquake may induce aseismic slip of the fault downdip of the main seismic rupture over timescales of days to years following the earthquake (e.g., Wang, 2004). We have approximated the integrated effects of shallow coseismic slip and deep post-seismic slip on the subduction fault by modelling the effects of coseismic and postseismic slip separately using the elastic dislocation model. During the coseismic phase, 20 m of slip is assumed in the thermal/ geodetic model seismogenic zone. Deep post-seismic slip is modelled by assuming 10 m of slip from the downdip limit of coseismic rupture to the halfway point of the coseismic transition zone. This "post-seismic rupture zone" is bounded on either end by transition zones, where slip linearly decreases from 10 m to zero; the updip transition zone extends to the deformation front and the downdip transition zone extends to the downdip limit of the coseismic transition zone. Both the amount of post-seismic slip and slip geometry are arbitrary, but reasonable values chosen to illustrate the effects of deep slip.

Figure 3 shows the modelled coseismic and post-seismic vertical deformation along the Vancouver Island profile. Due to the greater depth of the post-seismic slip compared to coseismic, the uplift/subsidence transition point is shifted landward. The sum of the coseismic and post-seismic deformation gives the total surface deformation once all earthquake-related slip has ceased (**Figure 3**). Deformation of the Earth's surface due to coseismic and post-seismic slip can be divided into three regions. Far inland from the deformation front, the surface will experience subsidence during both phases of slip (usually including the coastal region). Thus, the final amount of subsidence may be larger than predicted by the simple coseismic rupture model. Close to the deformation front, both phases produce surface uplift, resulting in a greater amount of total uplift than that produced coseismically. Between these two regions is an area that will first subside during coseismic rupture, and then uplift during post-seismic slip. This area is defined by the location of the uplift/subsidence transition points for coseismic and post-seismic rupture. This transition point lies slightly landward of the downdip limit of full rupture for each phase (~50 km for coseismic slip and ~75 km for post-seismic slip in the above model).

This simple model illustrates that post-seismic slip may significantly affect the vertical deformation at coastal sites. Post-seismic slip also could explain some of the discrepancies between the modelled coseismic deformation and the marsh data for Cascadia. The large observed subsidence between 140 and 170 km on the Columbia River profile may reflect deep slip of the fault. However, there is some uncertainty in the upstream data due to the effects of the Columbia River freshwater (e.g., references and discussion given by Leonard et al., 2004). The small amount of subsidence of the Vancouver Island coast may also reflect post-seismic slip. In this case, coastal Vancouver Island would have to lie in the "transition region", such that the final amount of subsidence is less than that from coseismic slip. Because these data are ~95 km from the deformation front, significant post-seismic slip would have to occur to a long distance downdip, i.e., 110 km or more from the deformation front (or to a depth of at least 30 km).

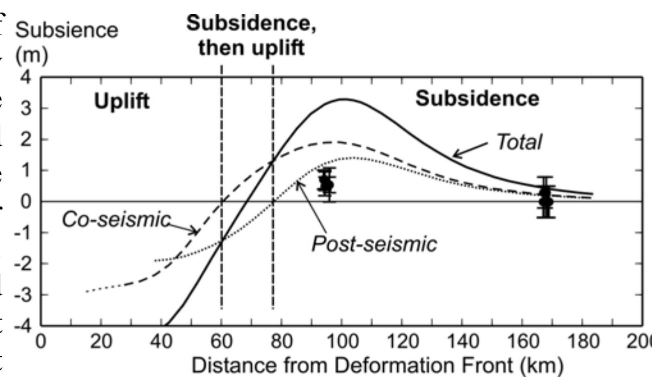

Figure 3. **Modelled elastic coseismic subsidence for the Vancouver Island profile (dashed line), subsidence for an arbitrary model of post-seismic slip on the deep subduction fault (dotted line), and the total amount of subsidence after both phases (solid line). With post-seismic slip, the motion of the ground surface can be divided into three regions: (1) an area with uplift during both phases, (2) an area that subsides during the earthquake, then uplifts during post-seismic slip, and (3) an area with subsidence during both phases.**

Viscoelastic relaxation. A large earthquake induces stresses in the surrounding material, which will relax over time if the temperatures are high enough for viscous behaviour. Elastic models provide a good approximation for coseismic deformation of the Earth's surface, but stress relaxation below the upper brittle-elastic layer following the earthquake requires a time-dependent viscoelastic model. Viscoelastic relaxation is difficult to model, as many of the parameters involved are not well-constrained. Although relaxation is generally thought to be most important on timescales of decades to centuries, Wang (2004) argues that viscoelasticity may explain short-term post-seismic transient deformation following great thrust earthquakes in NE Japan, SW Japan, Chile and Alaska. In particular, stress relaxation is used to explain geological and geodetic observations that show that the region of maximum coseismic subsidence became an area of rapid uplift within a few years after the 1964 Alaska and 1944-1946 Nankai (SW Japan) earthquakes (Wang, 2004 and references therein). If the Cascadia margin behaves in a similar manner, one may expect coseismic subsidence of the coast, followed by rapid uplift within a few years of the earthquake.

Due to post-seismic slip and viscoelastic relaxation, surface deformation following an earthquake is highly time-dependent. Although not well-constrained, both processes may result in a significant amount of uplift in a region that had experienced coseismic subsidence. It is so far not possible to infer this time-dependent deformation from the marsh data. The marsh data mainly constrain the elevation difference of the coast associated with great earthquakes but do not have adequate time resolution to detect post-earthquake motions of months to a few years. The subsidence recorded in marshes will represent the minimum amount of subsidence due to the Cascadia earthquakes, but the amount of coseismic subsidence may have been larger if either post-seismic slip or stress relaxation produced coastal uplift within the days or months following the earthquake.

5. Pre-1700 Cascadia Megathrust Earthquakes

Leonard et al. (2004) compiled estimates of coastal coseismic subsidence for the 1700 event, the last great earthquake to occur at Cascadia. This event affected the entire margin from northern California to central Vancouver Island, and was recorded at coastal marshes as a buried soil, often covered by a layer of sand inferred as a tsunami deposit. The date of this event has been determined accurately from tree-ring studies that constrain the death of trees submerged in the event to between the 1699 and 1700 growth seasons (e.g., Yamaguchi et al., 1997), and also correlating with the documented occurrence of a far-field tsunami in Japan in January 1700 (Satake et al., 1996).

Leonard et al. (2004) found that the pattern of 1700 coseismic subsidence was generally consistent with that predicted for the release of 500-800 years of strain accumulation using a simple elastic dislocation model. However, the 1700 event was only the latest of several earthquakes recorded by Cascadia coastal marshes; a sequence of up to 11 buried soils is present beneath the surface in some areas, representing the past several thousand years. Here, we assume that these buried soils represent subduction zone earthquakes (see discussion in Nelson et al., 1996) and use the information available from the marshes to examine how typical the 1700 great earthquake was of events over a greater time period.

Correlation of these older buried soils is often possible within individual estuaries, generally by "bar-code" matching (e.g., Atwater and Hemphill-Haley, 1997), but large uncertainties in radiocarbon dates make correlation over greater distances problematic. Therefore, we do not attempt to compile subsidence data on individual events, as done by Leonard et al. (2004) for the 1700 earthquake. Instead, we compile the available data for all previous events and estimate the mean and range of coseismic subsidence experienced by the various marshes over a number of earthquake cycles. For each location, we then compare this data with that for the 1700 event.

Coseismic subsidence is estimated here using published stratigraphic and sedimentological data, as described in detail by Leonard et al. (2004). The magnitude of coseismic subsidence is the difference in paleoelevation between each buried soil and the sediment immediately overlying it. Paleoelevation is estimated by comparing the properties of the buried sediment (organic content, macrofossils, and microfossil assemblages) with the sediment properties of modern intertidal elevational zones. Matching of the fossil sediment with a particular intertidal zone results in the assignment of a paleoelevation range. The error on the resultant coseismic subsidence estimates is primarily due to the width of the intertidal zones. In general, published data on pre-1700 buried soils is confined to organic content (e.g., peat, peaty mud) and sometimes macrofossils (e.g., Spruce, Triglochin), whereas many studies on the 1700 horizon also include detailed microfossil analysis, leading to higher precision estimates.

A table with references has been prepared (not included here) giving a summary of the compilation of coseismic subsidence estimates at each Cascadia marsh (see **Figure 4** for locations) both for the 1700 event and, where possible, the pre-1700 events. In each case, the range and mean of estimates is given, as well as information on the number of buried soils and estimates these represent. With the pre-1700 events, it is possible that some of the "buried soils" may represent either gradual sea level rise or localized faulting (e.g., Nelson et al., 1996), and thus the maximum number of soils may be an overestimate of how many megathrust earthquakes are actually recorded. The pre-1700 mean subsidence also may be biased towards larger events. Some events could be too small to be recorded, for example by occurring only a short time after the preceding event such that the marsh has undergone little interseismic uplift. Also, preliminary studies (e.g. Witter et al., 2003) suggest that some of the pre-1700 events ruptured only part of the margin and thus represent smaller

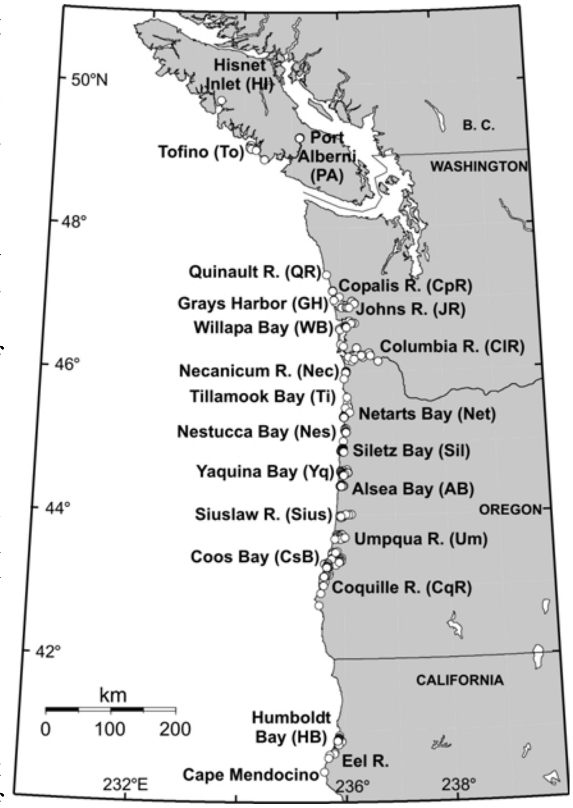

Figure 4. **Locations of sites with coastal subsidence data for past great earthquakes.**

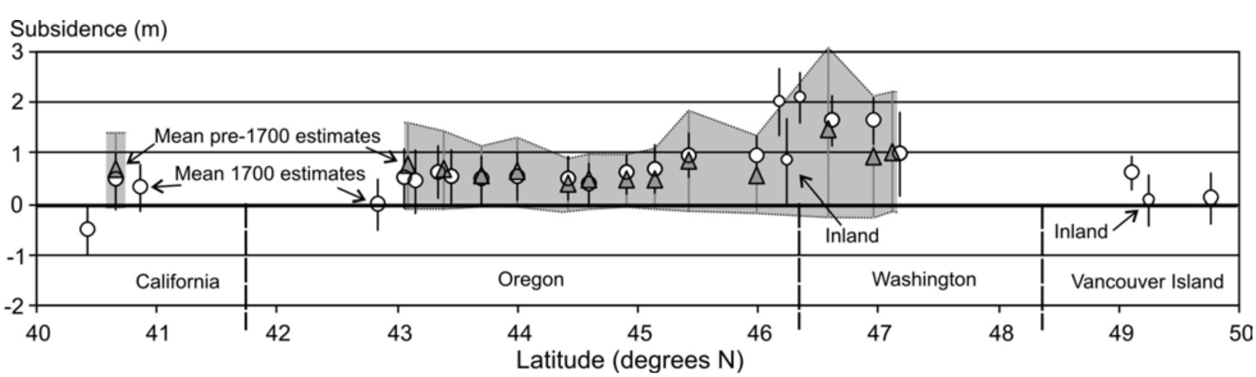

Figure 5. **Summary of site-mean coastal subsidence data with latitude for pre-1700 great earthquakes (and data range), compared to the 1700 earthquake subsidence.**

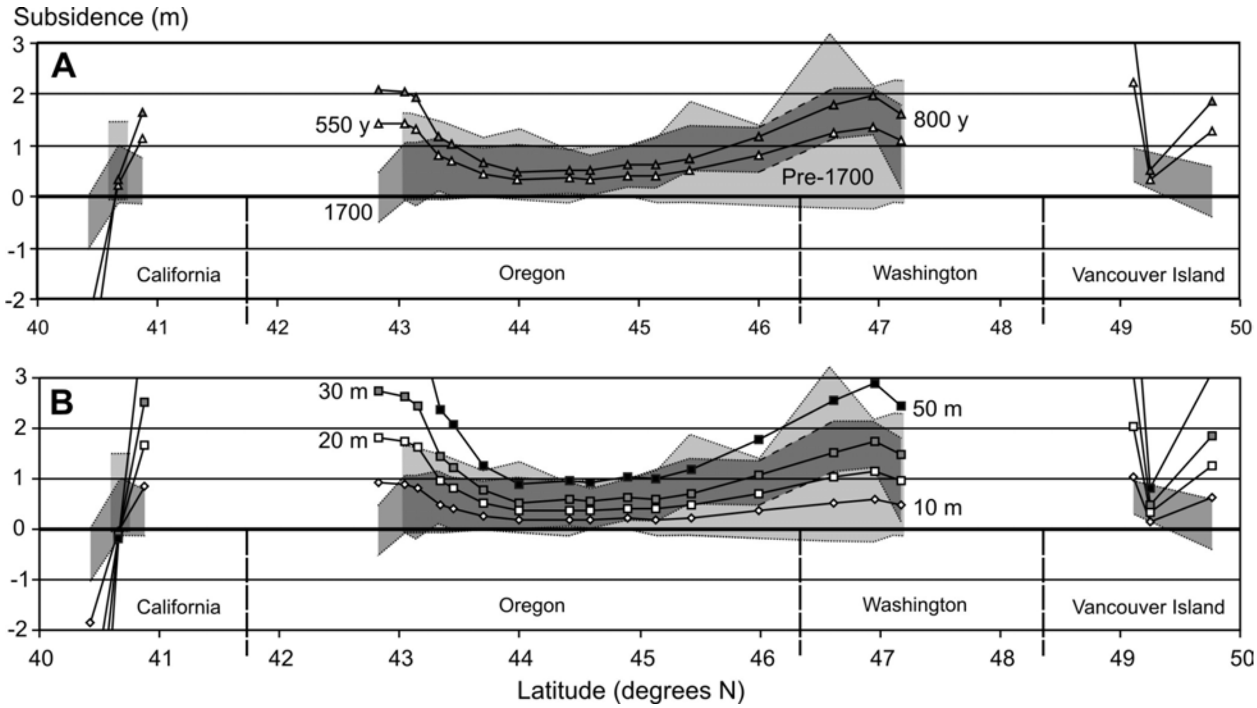

Figure 6. Comparison of subsidence for pre-1700 great earthquakes (light shading) and 1700 earthquake subsidence (dark shading), with elastic dislocation model predictions for several rupture displacements.

earthquakes than the ~ M_w 9, 1700 event; this question requires much more investigation.

The coastal results suggest that the 1700 earthquake is basically representative of a typical Cascadia megathrust event. However, in Washington and northernmost Oregon, somewhat greater subsidence occurred in 1700 than in most pre-1700 events, while in southern Oregon and northern California, 1700 subsidence was less than in most previous recorded events. At the southern end of the subduction zone, this may partially resolve the issue of the discrepancy between the dislocation model and the 1700 data, if the 1700 event released less strain there than most previous events.

6. Conclusions

For most of the Cascadia margin, Leonard et al. (2004) showed that coseismic subsidence inferred from coastal marsh data is in broad agreement with the elastic dislocation model of rupture, assuming strain release representing 500-800 yrs of convergence, and a rupture width consistent with the seismogenic zone inferred from geodetic and geothermal data. However, quite large variations in the rupture width do not strongly affect the predicted surface deformation at most marsh sites, as the data are significantly landward of the uplift/subsidence transition point. Humboldt Bay is the only area with data close to the transition point. These data may be better fit by using a slightly narrower rupture width than in the Leonard et al. (2004) and others' geodetically-constrained dislocation models.

The amount of subsidence at the north and south ends of the subduction zone is much less than predicted using an elastic model coseismic rupture with 500-800 yrs of strain release. This may be because: (1) the fault behaviour becomes more complicated near triple junctions than assumed in the dislocation model; (2) the 1700 earthquake did not release the full amount of accumulated strain; (3) the earthquake recurrence rate in these areas is higher than the rest of the margin; this is suggested for the southern region but there is no evidence that this is the case for the northern region; (4) the marsh data include rapid coastal uplift immediately following the earthquake due to either post-seismic slip along the deep part of the fault or viscoelastic stress relaxation.

Comparison of the coseismic subsidence recorded in 1700 with estimates for pre-1700 events suggests that the 1700 megathrust earthquake is generally representative of previous recorded subduction earthquakes. However, in Washington and northernmost Oregon, the 1700 subsidence was somewhat greater than in most previous events, while at the southern end of the subduction zone it was less than in most previous events.

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